The Distribution of Temperature and Wind Connected with Active Tropical Air in the Higher Troposphere and Some Remarks Concerning Clear Air Turbulence at High Altitude

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Abstract

A study is made of the distribution of wind and temperature in the higher troposphere and the lower stratosphere in cases with active Tropical air in middle latitudes. Only *observed* winds are used. A model for extending the fronts into the lower stratosphere is given.

In connection with the proposed model a case with observed clear air turbulence at high altitude is studied. Some remarks on the statistical treatment of the reported cases with clear air turbulence is made.

The concept of fronts has been used in synoptic meteorology for about 30 years as far as sea level charts are concerned. In synoptic aerology, however, the fronts have not been fully accepted, although J. Bjerknes, Palmén, Van Mieghem, and others already in the 30's proved that is was possible to use this concept also in the middle troposphere. The aerological work (in Europe) was founded, however, to a great extent on aircraft ascents, the ceiling of which was about 5 km. It was as a rule not difficult to identify the fronts up to this height, but when the soundings began to reach heights between 10 and 20 km new problems arose. It was firstly the swarm ascents, made possible by the Jaumotte meteorograph, and secondly the appearance of the radiosond that brought to the fore the problem of prolongation of the fronts into the higher troposphere (and the lower stratosphere).

It is especially when studying the three-dimensional structure of the atmosphere by means of vertical cross sections one meets this special problem.

The rapidly increasing importance of synoptic aerology during World War II, brought about especially by the development of aircraft efficiency, focussed the meteorologists' attention still more on conditions in the higher troposphere and the lower stratosphere.

This paper is a preliminary report of an investigation concerning the problem of extending the fronts into the lower stratosphere. This problem has been treated by J. BJERKNES and PALMÉN (1937), J. BJERKNES, MILDNER, PALMÉN, and WEICKMANN (1939), NYBERG and PALMÉN (1942), but the available material was mostly sparse, and wind observations from the critical region were entirely missing. The present author has taken up the problem because of the fact that the radiosond network nowadays by far exceeds that, on which the earlier works were founded.

[1] Written while on leave of absence from Swedish Meteorological and Hydrological Institute, Stockholm.

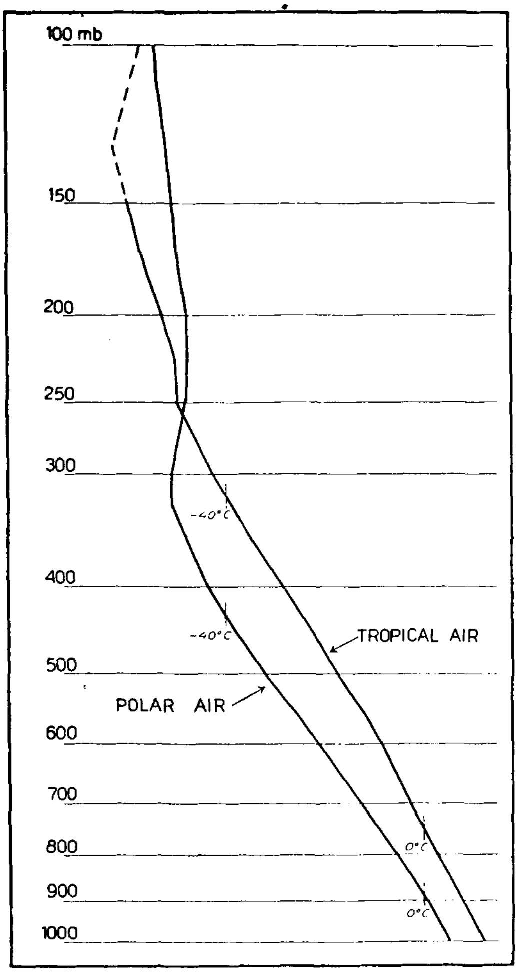

Fig. 1. Two soundings representative for Tropical and Polar air, respectively, in western Europe and over the middle Atlantic in November.

One week of November 1949 is treated by the author for another purpose. From this material two soundings were constructed[2], one representing Tropical air and the other Polar air (see fig. 1). The temperature difference between the two air masses increases with height to about 400 mb, from 7° C to about 16° C. Above the tropopause (at 325 mb) the Polar air is isothermal, and at 257 mb both air masses have the temperature of — 49.0° C. Then the horizontal temperature gradient is reversed and increases up to 130 mb, where the difference is about 11° C. Now, the question is whether this stratospheric temperature difference is concentrated to a front, as is the case in the troposphere, or not. Furthermore it is interesting to study the behaviour of the front in the upper part of the troposphere.

In order to study these phenomena I have chosen two cases, one from November 1949, and another from October 1949.

On the 7th of November 1949, an almost stationary front was located over western Europe, extending towards westnorthwest to the middle of the Atlantic. This front began to move northeastwards over the Atlantic, later towards eastnortheast. A rather strong sea-level low developed during this period, and on the 9th at 0300 GCT its warm front had just reached northern Ireland and western France (see fig. 2 a). The cold front extended from westernmost Ireland down to the Azores.

A high over Russia was almost stationary, and the southeasterly to southerly air current at its southwestern and western side was very stable and had persisted for about four days.

The cold air over Central Europe between these two warm currents was maritime Polar air, which had come from the Atlantic and which had partly been Arctic air.

On the 500 mb chart (fig. 2 b) the advection of the Tropical air over and east of the British Isles is very pronounced. The warm front is especially well-defined over northwestern Germany, the Netherlands, and Belgium, and here the density of the radiosond network is so great that the analysis may be considered as quite reliable.

The broad southerly current over eastern Europe is well developed, but the front at 500 mb connected with this current is not as strong as the western one.

As a whole the sea level chart as well as the 500 mb chart in fig. 2 gives a good picture of an upper-air pattern, which is rather common, at least if western and eastern Europe are considered separately.

In order to illustrate the distribution of wind and temperature in this case, I have drawn a vertical cross section parallel to the latitude circle of about 54.5° N. The section extends from Valentia in western Ireland to the British Zone in Germany. The cross section can not be extended further eastwards, because the

[2] The details of the computations will be accounted for in the complete report, which will be published in *Arkiv för Geofysik*, vol. 2 (in print).

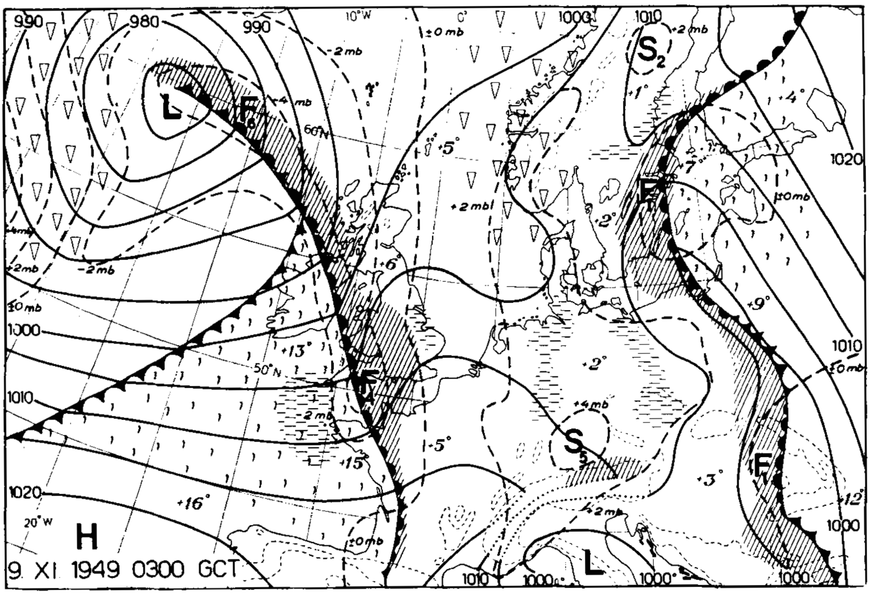

Fig. 2 a. Sea-level chart for 0300 GCT 9 November 1949.

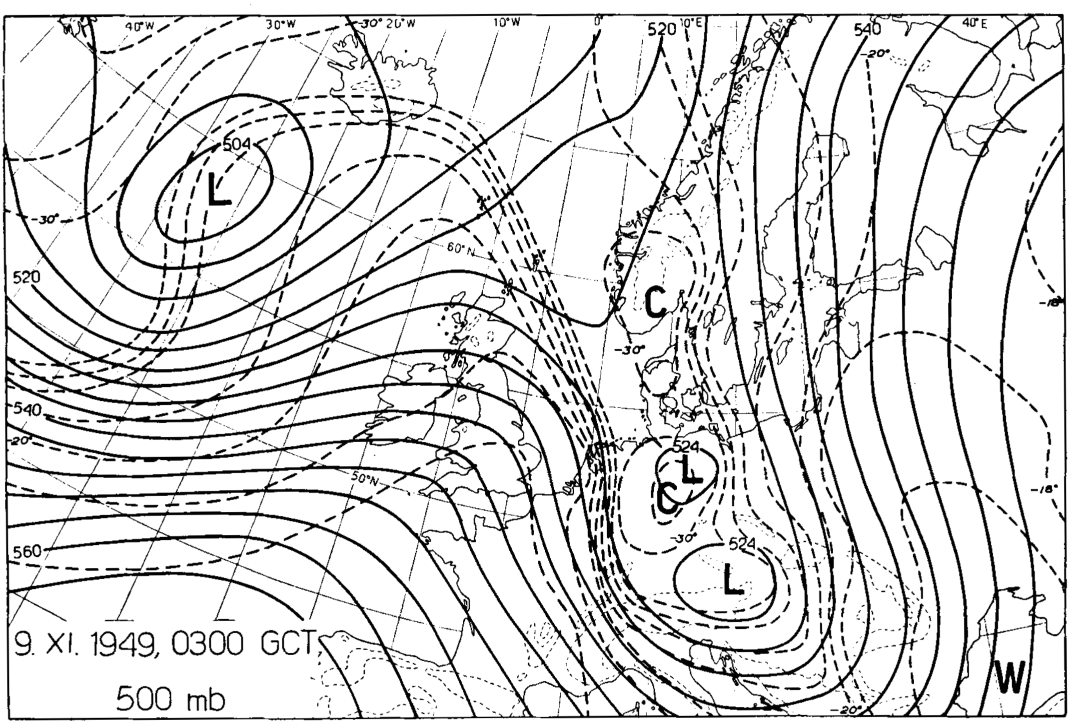

Fig. 2 b. 500 mb chart for 0300 GCT 9 November 1949. Solid lines are isohypses for every 4 geodynamic decameters, dashed lines are isotherms for every 2° C.

Fig. 3 a

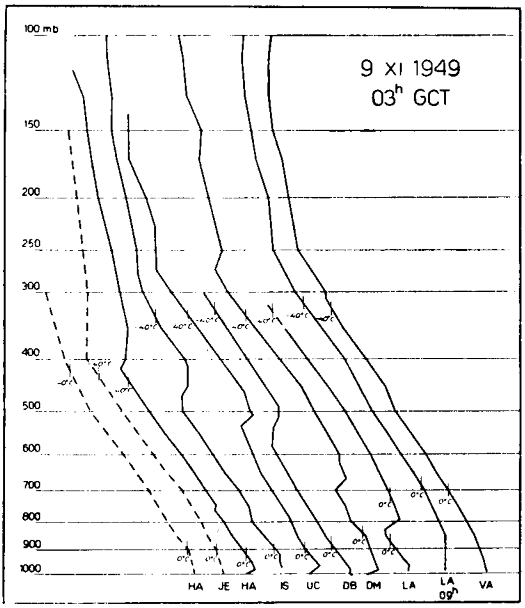

Fig. 3 a. Ascent curves, 0200 or 0300 GCT 9 November 1949, for Hannover and Iserlohn (British Zone in Germany), Uccle (Belgium), De Bilt (the Netherlands), Downham Market, Larkhill, and Valentia (British Isles). The left curves are the ascent curves for Hannover and Jever (British Zone in Germany) for 2000 GCT 8 November, and the second from the right is the 0900 GCT ascent from Larkhill on the 9th of November.

Fig. 3 b. Vertical cross section from Valentia (Ireland) to Hannover (British Zone in Germany) for 0300 GCT 9 November 1949, approximately along 54.5°N. Thin dashed lines are isotherms (° C), thin solid lines isovels for *observed* wind speed (m sec⁻¹, projected on a direction of 330°). Intermediate values of temperature and wind speed are given by dotted and dashdotted lines, respectively. Heavy lines represent tropopauses, frontal boundaries, and inversions. Thin vertical lines give the location and vertical extent of temperature soundings.

Fig. 3 c (next page). Vertical cross section as in fig. 3 b, giving isentropes (thin dashed lines, °K) and isovels for *observed* wind speed (thin solid lines, m sec⁻¹). Intermediate values for the wind speed are given by dashdotted lines. Frontal boundaries, tropopauses, and inversions given as heavy lines. Thin vertical lines give the location and vertical extent of the wind soundings.

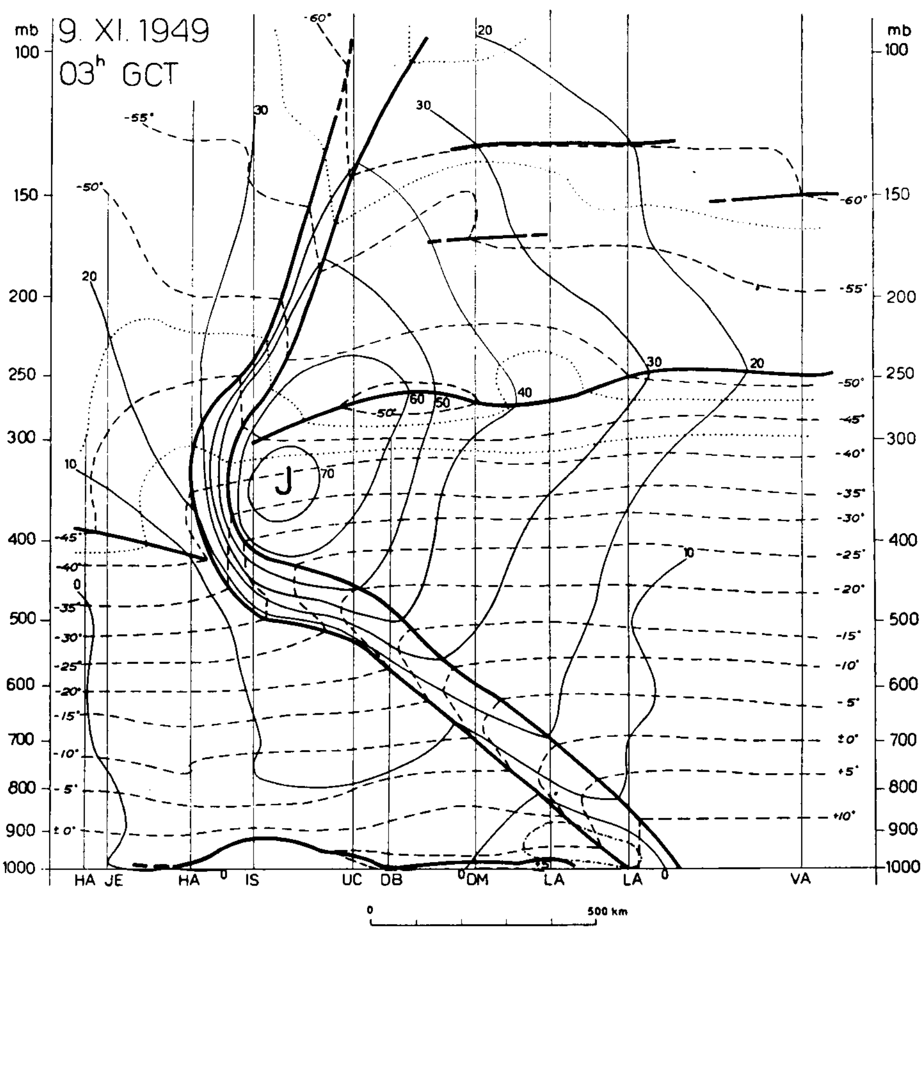

Fig. 3 b

frontal system of eastern Europe would then enter the picture. To avoid this difficulty we have to some extent changed our space cross section to a time cross section in the easternmost parts. The changes within the Polar air during six hours are quite small and it is therefore permissible to use the Hannover and Jever soundings from the 8th of November, 2000 GCT, in our section. The ascents are plotted so as to be in the right positions vis-à-vis the front, the velocity of which parallel to the cross section was observed to be about 35 km h^{-1}.

Furthermore the sounding from Penzance at 0300 GCT on the 9th of November is a little too warm in the higher troposphere. This fact has been revealed through the analysis of the upper-air charts, and we have therefore substituted the Penzance ascent by the one from Larkhill, at 0900 GCT on the same day. If we use the estimated velocity of the front the latter ascent will be situated approximately at the right place.

The ascent curves of this cross section are given in fig. 3 a. The Polar front is easily seen in all soundings except those three to the left. The inversion in the Hannover ascent from 0300 GCT may be taken as an indication of the front, but the earlier history of this air as well as the wind distribution given by this ascent, does not support this solution. This question will be subject for a more detailed discussion later in this paper.

The tropospheric front is very distinct in the isotherm field (see fig. 3 b). It is a particularly interesting fact that practically all soundings give inversion in the lower portion of the frontal layer and weak vertical temperature gradient in the upper portion. In the fully developed stage almost all soundings through this warm front give the same picture. Studying the ascents

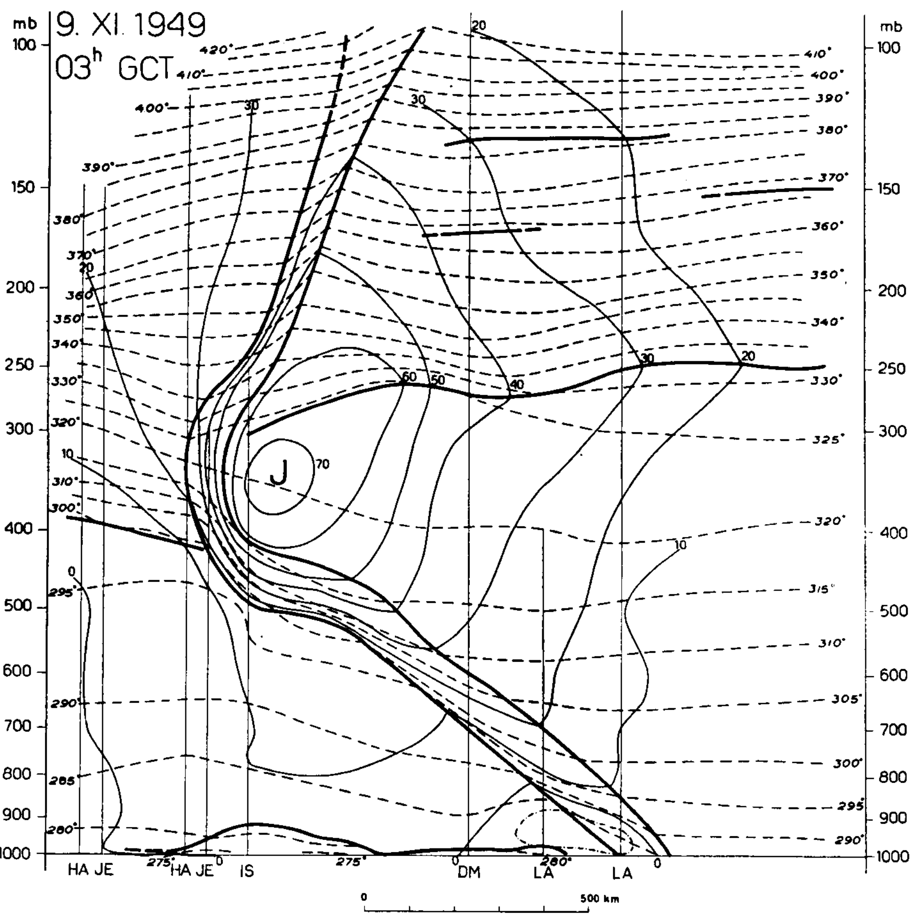

Fig. 3 c

from the 8th through the same warm front one comes to the conclusion that this fact may be taken as an indication that the vertical motions on one side of the front are rather strong. A discussion of this interesting question would, however, lead too far.

The tropospheric Tropical air is nearly barotropic, except in the neighbourhood of the front and in the westernmost part, where the prefrontal cooling ahead of the cold front can be seen. The Polar air is slightly barocline in the troposphere and near the upper parts of the tropospheric front it is quite strongly barocline.

In the stratosphere the isobaric temperature differences within the Polar and the Tropical air, respectively, are rather small, except in the Tropical air near the front. The contrast between the two air masses amounts to about 5° C and seems to be concentrated to the front.

Fig. 1 showed the well-known fact that the Tropical air is colder than the Polar air above, say, 350—250 mb. If the Tropical air is active throughout the troposphere and lower stratosphere, then the Polar front must act as a warm front in the troposphere and as a cold front above the lower tropopause in the Tropical air.

Throughout this paper we have used the expression 'stratospheric Tropical air' and the like. According to our opinion it is not correct to regard the tropopause as the upper limit of the Tropical and Polar air (in middle and high latitude). *The tropopause (or tropopauses) is (are) mostly a phenomenon (phenomena) occurring within the air mass in question.*

The wind field (figs. 3 b and 3 c) contains a strong jet at about the 350 mb level with a wind speed of about 75 m sec^{-1}. The figures give the component of the wind speed parallel to 330°, *i.e.* a direction almost perpendicular to the cross section. The vertical wind shear through the tropospheric front is about 13 m sec^{-1} km^{-1} in the lower part and about 33 m sec^{-1}km^{-1} between the 400 and 500 mb levels.

The isobaric wind shear is considerable. The anticyclonic shear is about 9 m sec^{-1} (100 km)$^{-1}$, if we assume a linear fall of the wind speed from the jet axis to the station Downham Market (DM). The real shear is, however, probably greater in a region somewhere in between, where values approaching the Coriolis parameter are likely.

The cyclonic wind shear is definitely greater. Assuming a linear velocity field between the stations Iserlohn (IS) and Hannover (HA) we get 32 m sec^{-1} (100 km)$^{-1}$. Comparing the velocity distribution both in horizontal and vertical direction where the Iserlohn ascent passes through the warm front, we find justification for concentrating most of the isovels to the front. In this manner we obtain an isobaric cyclonic wind shear of about 60 m sec^{-1} (100 km)$^{-1}$, or more than five times the Coriolis parameter.

There is, unfortunately, one complication that ought to be discussed a little. The launching time of the ascents may deviate somewhat from what is indicated in the synoptic reports, and the observed wind speed has both systematic and accidental errors. Assuming a time error of 30 minutes and a speed error of 10 m sec^{-1} and using the estimated velocity of the front, we obtain 22 m sec^{-1} (100 km)$^{-1}$ as value for the cyclonic wind shear, with a linear fall of the wind speed. The assumed errors are certainly too great in most cases and therefore we may give the interval 45—60 m sec^{-1} (100 km)$^{-1}$ for the wind shear as a convenient one in this case.

The isobaric wind shear is thus greatest in the region where the temperature difference disappears and the front becomes vertical. Thus, with the aid of the wind observations it should be possible to trace the front between 400 and 250 mb. In this region it is often difficult to analyse with the aid of temperature alone, because the temperature distribution is blurred by strong vertical motions. (In the complete report we are going to discuss to some extent the vertical motions connected with this model.)

Here it must be emphasized that throughout this paper we use the ordinary definition of a front as a discontinuity in density. It is only in the region where the temperature difference between the two air masses disappears that we use a special definition of the front. In this certainly rather thin layer we define the front as the region with great horizontal cyclonic shear.

As item two we have chosen a case from October 1949. During the period 22nd to 24th unstable, fresh maritime Polar air covered the greater part of northwestern Europe. Between Greenland and the Scandinavian peninsula there was a continuous northerly flow of Arctic air, which was transformed on its way southwards over the ocean. It was therefore not possible to trace an Arctic front in this region. Over the Atlantic Ocean south of 50 °N a series of disturbances moved eastwards to the southern parts of the British Isles and France. One of these systems which on the morning of October 25th was situated south and southwest of the British Isles, became more intense than the preceding ones. The warm front of this system was very distinct and can easily be found on the sea-level chart. (The sea-level fronts are indicated in fig. 4 with the international symbols).

The 500 mb chart (fig. 4) shows a strong southwesterly current over southwestern England and adjacent seas. This current veers west and passes over northern Germany, for instance, as a due westerly wind. The isotherms are crowded where the Polar front intersects the 500 mb level, and the temperature difference between the Tropical and the Polar air through the front is about 10° C.

This time I prefer to show a time cross section instead of a space cross section. The British radiosond network, with its six hours intervals, is well fitted for this purpose. In fig. 5 b I have given a time cross section for Aldergrove, the radiosond station in northern Ireland, and for the period 0300 GCT 24th—0900 GCT 25th of October 1949.

The ascent curves given in fig. 5 a have essentially the same characteristic features as have those of fig. 3 a. The warm front is easily detectable from 1500 GCT on the 24th and thereafter. The deep layer between 418 and 330 mb in the 0900 GCT ascent is not be regarded as the frontal layer. The lowest characteristic point (418 mb) must be identified with the lower tropopause of the soundings six hours earlier.

The isotherm field of fig. 5 b gives as a whole the same picture as the one in fig. 3 b. As in the first case, the difference in vertical stability on both sides of the front between, say, 350 mb and 200 mb is very striking. Comparing the ascents from 0900 GCT and 2100 GCT on the 24th, we get the following figures for the vertical gradient of temperature and potential temperature. The ascents give (the figures for Tropical air are given within brackets) a temperature difference of — 8.9° C (— 23.9° C) and a potential temperature difference of + 41.0° C (+ 18.5° C). The mean

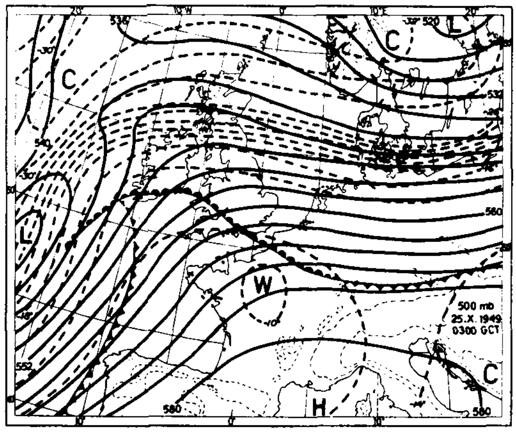

Fig. 4. 500 mb chart for 0300 GCT 25 October 1949. Isohypses are given as solid lines for every 4 geopotential decameters, isotherms as dashed lines and for every 2° C. The fronts at the sea-level are given with international symbols.

vertical distance between the 200 and the 350 mb levels is 3,713 m. Thus, we get as values for the vertical gradient of temperature and potential temperature — 2.4° C km^{-1} (— 6.4° C km^{-1}) and + 11.0° C km^{-1} (+ 5.0° C km^{-1}), respectively.

The difference in wind velocity on both sides of the front between 300 and 250 mb is great. In the Polar air just ahead of the front, there is a wind speed of about 40 m sec^{-1}. In the Tropical air on the other hand the jet seems to reach a value of at least 80 m sec^{-1}.

If we now take into consideration (1) the great differences in vertical stability, (2) the observed wind speed on both sides of the front and (3) the difference in temperature between the two air masses, measured along an isobar, below and above the height of zero temperature difference (in fig. 3 b at about 325 mb and in fig. 5 b at about 310 mb) we reach the following conclusion. *There is no reason not to extend the tropospheric warm front into the lower stratosphere, where it acts as a cold front. Between about 350 and 200 mb the front is a boundary between tropospheric Tropical air and stratospheric Polar air.*

Remarks on clear air turbulence

Finally I want to discuss a special phenomenon, which in many cases seems to be connected in one way or another with the atmos-

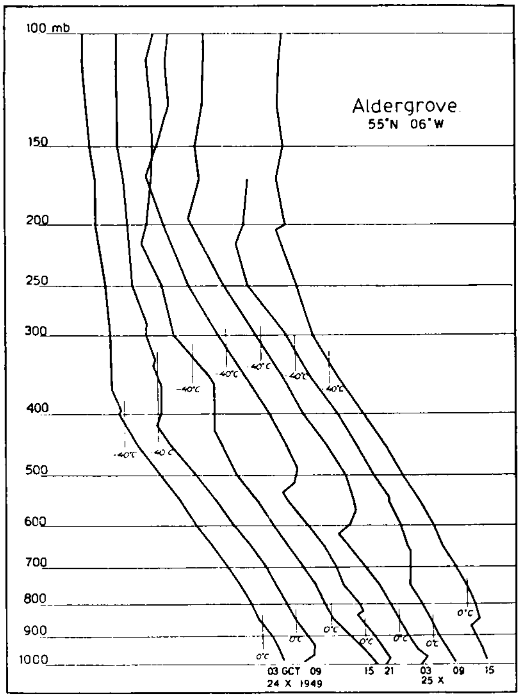

Fig. 5 a. Ascent curves for the British radiosond station Aldergrove, northern Ireland, for the period 0300 GCT 24 October—1500 GCT 25 October 1949.

pheric model discussed above. During the last ten years clear air turbulence above 5 km has been subject of investigations, especially in Great Britain and U.S.A. We shall start with a brief summary of two papers, one by HISLOP (1951) and the other by BANNON (1951), both summarizing the British investigations.

A series of experimental flights were made by British aeroplanes between 6 and 12 km above sea level in order to try to get quantitative data of the clear air turbulence.[1] This type of turbulence[2] was not infrequently observed over the British Isles above a height of 6 km. The turbulent layers were usually rather shallow; out of 290 cases 83 % were observed in layers of less than 900 meters depth, but in 8 % of the cases the layer was reported to be more than 1,800 meters thick.

The distribution with height shows no great variation (58 cases in the layer between 6 and 7.2 km, 72 between 8.4 and 9.6 km, and 55 between 10.8 and 12 km). If the material is grouped with respect to the height relative to the tropopause, then we get a significant decrease of frequency above the tropopause level.

The jet streams and their margins are the site of many of the reported turbulence cases. Bannon tries to combine the Richardson num-

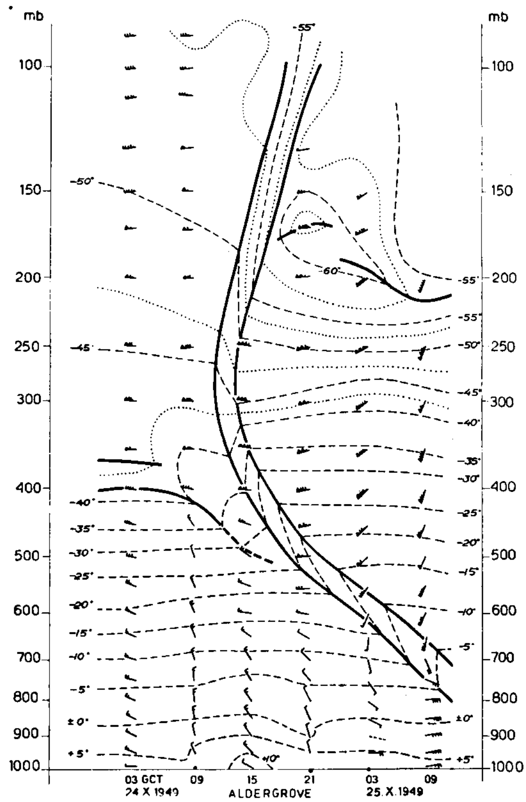

Fig. 5 b. Time cross section for Aldergrove for the period 0300 GCT 24 October—0900 GCT 25 October 1949. Isotherms are given as dashed lines (°C), and *observed* winds are plotted in the conventional manner. Full barb 5 m sec^{-1}, half barb 2.5 m sec^{-1}, triangular barb 25 m sec^{-1}. Intermediate values of temperature are given as dotted lines. Heavy lines are frontal boundaries and tropopauses.

[1] In these investigations only gusts giving a vertical acceleration ≥ 0.2 g are considered. The type of aircraft used determines to some extent what kind of turbulence element will be found in the atmosphere. Spitfires and Mosquitos were used and the turbulence eddies causing bumps have the dimensions of 15 to 500 meters according to Bannon.

[2] In the British investigations, which were devoted to frictional or dynamic turbulence above 6 km, the thermally induced clear air turbulence in lower layers is disregarded.

ber[1] and the occurrence of turbulence. He comes to the conclusion that the bumpiness is related to the Richardson number, but the correlation is not too good. The same condition seems to be valid also for the horizontal wind shear[2] contra the frequency of clear air turbulence. Bannon also concludes that these two possible causes of turbulence are to some extent independent.

In an unpublished report by HARRISON (1950), quoted by HISLOP, the suggestion is made that the wind shear in the horizontal is a significant factor. This suggestion is founded upon experiences from United Airlines flights with DC-6's between 4.5 and 7.5 km height in U.S.A.

In Bannon's paper nothing specific is said about the connection between clear air turbulence and the fronts. He reviews a few examples of flights during which bumpiness was encountered in inversions or isothermal layers.

After this short review of the British work in this field a synoptic example will be given. On April 25th 1949, at 1200 and 1220 GCT an aeroplane encountered clear air turbulence over southeast England. The cross section in fig. 6 from Aldergrove in northern Ireland to Downham Market in southeast England was located quite near the critical region. As there are no radiosond measurements available for 1200 GCT, the 0900 GCT soundings are chosen. This cross section gives the same picture as do the earlier cross sections (*cf.* figs. 3 b and c, figs. 5 b and c)[3]. The front is easily traceable and the temperature difference between the two air masses is of the order 10° C. The cold air in the troposphere is almost barotropic, except near the frontal boundary, and its tropopause is well defined.

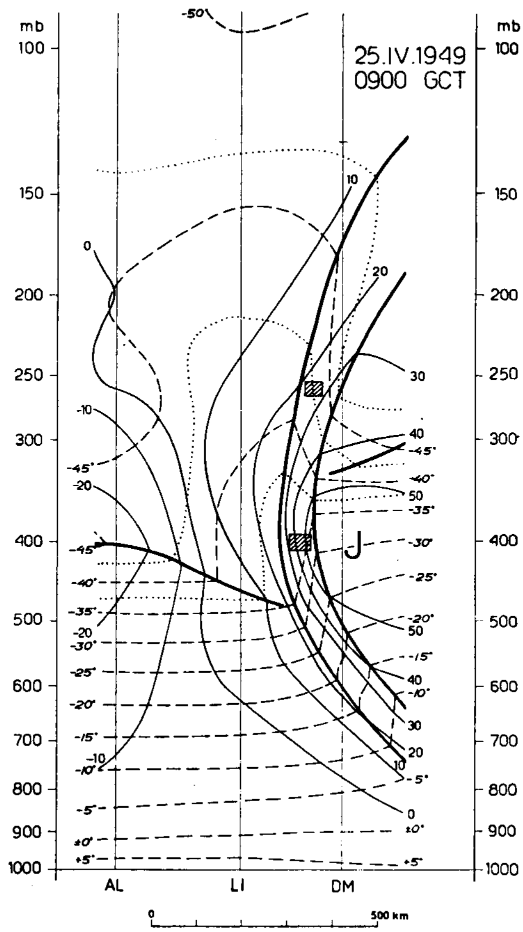

Fig. 6. Vertical cross section Aldergrove-Liverpool-Downham Market, 0900 GCT 25 April 1949. Heavy lines represent tropopauses and frontal boundaries, thin dashed lines are isotherms (° C), and thin solid lines isovels for *observed* wind speed (projected on a direction of 190°, m sec⁻¹). Intermediate values of temperature are given as dotted lines. Hatched areas indicate the regions where clear air turbulence was observed by special search aircraft.

The maximum horizontal shear through the front is about 60 m sec^{-1} (100 km)$^{-1}$, about the same as in the case given in fig. 3.

The regions within which clear air turbulence was reported by an aircraft belonging to the British project, are indicated by hatched areas on the cross section in fig. 6. The right position vis-à-vis the front has been obtained through an estimate of the velocity of propagation of the front, obtained from consecutive sections and upper air charts. The areas

[1] The Richardson number

$$R_i = \frac{g}{T}\left(\Gamma + \frac{\partial T}{\partial z}\right) \bigg/ \left(\frac{\partial v}{\partial z}\right)^2 \qquad (1)$$

(where g is the gravity acceleration, T the absolute temperature, Γ the dry adiabatic lapse rate, v the horizontal wind, and z altitude, measured upwards) is widely used in the theory of turbulence.

[2] Bannon's definition of the horizontal shear differs from the ordinary one, but this fact does not change the qualitative reasoning.

[3] The fact that this front acts as a cold front in the troposphere does not alter the reasoning in the discussion of the earlier cases. In the complete report we shall study all types of fronts in connection with the proposed frontal model.

covered by the hatching are 50 km wide and about 300 m deep, and may be considered to give the true position of the turbulent area relative to the front.

It is obvious that the lowest area coincides almost completely with the region of maximum horizontal wind shear.

The depth of the turbulent layers is a very interesting question. If turbulence were encountered in connection with the frontal model given in this paper, it is reasonable to assume that the deepest turbulent layers would be observed at the height where the front changes character (in fig. 6 at about 400 mb). A careful investigation whether this really is the case cannot, however, be done by the present author, because of the fact that he does not have the material at his disposal.

Considering the importance of the horizontal contra the vertical shear of the wind, the following statement can be made. In Polar fronts as those in figs. 3 and 6, the isovels in the cross sections are almost parallel to the frontal boundaries. This is true in the great majority of cases as synoptic investigations show (see *e.g.* PALMÉN and NEWTON, 1950, fig. 3).

Assuming the thermal wind equation to be valid in the following form (see HEWSON and LONGLEY, 1944, p. 99—101)

$$v = \frac{g}{fT}\,(z - z_0)\,\frac{\partial T}{\partial x} \qquad (2)$$

(where the symbols have the same meaning as in formula 1), we get a simple figure for the vertical wind shear. If we put $g = 9.81$ m sec^{-2}, $f = 10^{-4}$ sec^{-1}, $z—z_0 = 1$ km, $\frac{\partial T}{\partial x} = 1°$ C (100 km)$^{-1}$, and $T = 250°$ K, $v \sim 4$ m sec^{-1} km^{-1}.Thus, for every °C per 100 km of the horizontal temperature gradient the geostrophic wind increases with height with 4 m sec^{-1} km^{-1}.

Assuming a front of 1 km vertical depth and with an inclination of 1/100, we get with the above-mentioned assumptions a horizontal (thermal) wind shear of about 4 m sec^{-1} (100 km)$^{-1}$. Thus, the often observed temperature difference 10° C through the front gives a horizontal (thermal) shear of about 40 m sec^{-1} (100 km)$^{-1}$, through the *sloping* front. This is in agreement with synoptically observed values of this quantity, if we study the upper part of the troposphere. In the above discussion we have disregarded the conditions in the lower part of the troposphere. The real horizontal wind shear is slightly modified by the wind distribution in the lower layers, but the given order of magnitude of the wind shear is correct. Here it must be emphasized that with the network of wind-observing stations of to-day, the distance between the stations being at best about three times the horizontal width of the front, it is dangerous to assume linear distribution of wind velocity.

If the above-mentioned assumptions are fulfilled, it is clear that strong vertical wind shear is connected with strong shear in the horizontal in the case of a *normally sloping* front, if the isovels are (nearly) parallel to the frontal boundaries.

The vertical wind shear vanishes in the region where the temperature difference between the two air masses disappears, in fig. 6 at about 360 mb, and where consequently the front becomes vertical.

Before giving a short discussion of the statistical treatment of the reports of clear air turbulence we want to make two assumptions.

Firstly we assume that a great part of the cases with clear air gusts is connected with more or less distinct fronts, interpreted as in this paper, and the reported cases published up till now give the reader this impression. Having no opportunity to use the detailed reports from the search flights, we must consider this connection between fronts and clear air turbulence as a working hypothesis.

In connection with this assumption a remark may be made concerning fig. 10 in Hislop's paper. This figure gives an exemple of a jet stream over western Europe, and the region where clear air turbulence was observed is indicated. From the figure the reader gets the impression that the turbulent area is connected with *anticyclonic* wind shear. The fact is, however, that the chart is a copy of the 300 mb chart, given in the *London Daily Weather Reports, Upper Air Section*, for 0300 GCT. The turbulence was observed at 1300 GCT according to table I in Hislop's paper. Furthermore the mean height of the 300 mb surface over the British Isles was 8760 m, whereas the turbulence was encountered between 5 and 8.6 km. *As a matter of fact cyclonic wind shear,* *and not anticyclonic, existed in the critical region at about 1300 GCT.*

Secondly we assume that the frontal model given in this paper is not too far from truth, and this opinion seams to be supported by the examples put forth in this paper as well as in the complete report.

If these two conditions are fulfilled, a sampling of gust cases from all levels between 5 km and 12 km cannot be examined statistically without a special technique. Any crude method will certainly give disappointing results. This is obvious from what is said above and from the temperature and wind distributions given in figs. 3, 5 and 6. The correlation between horizontal wind shear, vertical wind shear, horizontal and vertical temperature gradient on one hand and clear air turbulence on the other cannot be studied when all reports are taken by the bulk.

If clear air turbulence were reported above 700 mb in connection with the front in fig. 3, one should distinguish between four different cases. (1) Between 700 and 500 mb the (cyclonic) horizontal and vertical wind shear are rather considerable. (2) In the layer at about the 500 mb level the inclination of the front is small and consequently the horizontal wind shear is also small but the vertical shear great. (3) A report from the region where the front is vertical (as in fig. 6, lower hatched area) will correspond to zero vertical wind shear and maximal horizontal (cyclonic) wind shear (4) Higher up the vertical wind shear is increasing but with the opposite sign compared with tropospheric values.

Thus, it seems likely that any theory which tries to explain the occurrence of clear air turbulence must take all these different cases in account. The purpose of the latter part of this paper is to give a hint to a possible way of attacking the problem and to give some values of the involved parameters.

Finally it should be pointed out that we might here have a possible way of attacking the prognostic problem of the clear air turbulence.

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